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EARLY THERMAL HISTORY OF THE TERRESTRIAL PLANETS

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August, 1966

Contract No. NSR-24-005-047

Prepared by

UNIVERSITY OF MINNESOTA
Minneapolis, Minnesota

For

HEADQUARTERS, NATIONAL AERONAUTICS & SPACE ADMINISTRATION
Washington, D. C. 20546

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EARLY THERMAL HISTORY OF THE TERRESTRIAL PLANETS

ABSTRACT

Using a homogenized earth as a prototype planetary model and using new estimates of the earth's age and average radioactivity, it is possible to determine the thermal evolution of various sized protoplanetary spheres up to the melting point. If the melting point of silicates and iron is exceeded in the lifetime of the planet, it is assumed that the planet will differentiate a core and a crust.

The protoplanetary models considered have the same composition as the present earth and therefore contain free iron dispersed throughout their interiors. There exists a maximum mass planet, intermediate to the Earth and Mars in size, which can exist without differentiation. If Mars has the same composition as the Earth, or the same composition as a completely oxidized Earth, it will not differentiate during its lifetime. Therefore, the thermal calculations for Mars are consistent with the lack of differentiation which is implied by the moment of interia, the lack of magnetic field and radiation belts, surface coloration and lack of tectonic surface features.

If the Moon is a silicate body having the composition of the Earth's mantle it exceeded the melting point of silicates early in its history and is probably a differentiated body.

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EARLY THERMAL HISTORY OF THE TERRESTRIAL PLANETS

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INTRODUCTION

There is much evidence that the Earth is a chemically differentiated body. If we assume that this was not the original state of affairs, then the differentiation of the original, presumed homogeneous body, into a zoned body consisting of a crust, mantle, and core is one of the most significant event or sequence of events that has occurred in the evolution of the Earth. A favored current hypothesis is that the planets accreted from relatively cold solid particles. The mechanics of the accretion process, the size and temperature of the accreting particles and the rate of accretion are speculative and this limits our ability to define the initial conditions of the fully accreted planet. Estimates of the initial near surface temperatures at the end of the major portion of the accretion stage of lunar to Earth sized objects vary from about 100 to 1000^{o}K. The temperature profile in a primitive planet will be governed by the rate at which mass, and therefore, gravitational energy is added, the rate at which it can be radiated away from the surface, and the initial temperature of the accreting particles and their change of temperature during the accumulation process.

The subsequent thermal history of a fully accreted body is controlled by the heating caused by the decay of radioactive isotopes and the redistribution of heat internally due to lattice conductivity, and, at higher temperatures, radiative transfer. In the early stages heat is being generated more rapidly than it is being removed and the body gradually warms up. The onset of partial melting and the subsequent possibility of rapid redistribution of heat and mass by material transport, which we shall call differentiation, signals the end of the stage which we call the "Early Thermal History." The latent heats associated with partial melting, the convective terms in the heat transport equation, convective heating due to material transport, gravitational energy made available in the process of mass redistribution and the redistribution of radioactive sources must be taken into account in the further thermal history of the planet. We shall consider only the early thermal history of bodies having the composition and mass of the terrestrial type planets.

Even ignoring the complications of the later thermal history, the uncertainties of the initial conditions, composition and thermal properties tend to make thermal history calculations suggestive rather than definitive. It is encouraging that previous thermal history calculations are broadly consistent with the known facts about the earth's interior. The melting point of iron is exceeded at depth in the mantle early in the history of an earth-sized planet and stays above the melting point at the present time. This is consistent with the formation of a core and a presently liquid core. The eutectic melting point of silicates is reached in the most realistic models but the latent heat of fusion and the efficient removal of heat by penetrative convection upward would tend to stabilize the system at a temperature below the complete melting point. Even if the considerable gravitational heating due to core settling causes complete melting, convection and redistribution of radioactivity would occur and it would be difficult to maintain temperatures in the mantle much above the melting point of the minimum melting phase. Although the seismic data indicates that the entire mantle transmits shear waves, this is not inconsistent with partial melting.

Thermal history calculations also show that the upper part of the mantle has the highest geothermal gradient and is closer to the melting points of silicates than other regions of the Earth. This is consistent with the presence of a low-seismic velocity zone in the upper mantle and higher seismic attenuation and electrical conductivity.

The astronomical data regarding Mars suggests that it is a relatively homogeneous body having roughly the same composition as the Earth. By homogeneous we mean that the density varies with depth primarily due to self-compression and there is no large concentration of mass toward the center such as exists in the Earth in the form of a heavy core. The presence of a thick atmosphere, volcanism, mountains, and a magnetic field on the Earth are all indirectly related to the fact that the Earth is a differentiated body and similar features are also lacking on Mars. The question arises, can we explain these profound differences between Earth and Mars simply by the difference in size of otherwise initially identical bodies or must we resort to more drastic measures?

Previous thermal history calculations which seemed satisfactory in their predictions regarding partial melting and hence differentiation for the Earth also lead to extensive melting for Mars even for very low initial temperatures (ref. 1). These calculations assumed chondritic compositions for both the Earth and Mars, and MacDonald (ref. 1) concluded that those bodies must differ significantly in their radioactive concentrations, a chondritic composition being appropriate for the Earth but not for Mars.

More recent studies suggest that the chondritic analogy is not appropriate for the Earth, particularly in regard to the radioactive abundances. This reopens the question. It is not clear that it will solve it since the uranium content in the mantle required to give the present observed heat flow in the Earth is about 3 times the uranium content required in a chondritic Earth. The terrestrial K:U ratio, however, is about one-eighth the chondritic ratio and because of the shorter half-life of K relative to U and Th, will make the heating up of a planet more uniform in time. Previous thermal history calculations of Earth and Mars are also not directly comparable since the Earth model is already differentiated at the start of the calculation, i.e. a core is already present and the radioactivity is already concentrated toward the surface of the Earth.

Neither an optical nor a dynamic flattening are available for Venus, and thermal calculations provide our only clue regarding conditions in the interior of this planet. Because of the similarity between the Earth and Venus in mean uncompressed density and total mass, one would expect conditions in the interior to be similar. The thick atmosphere on Venus, most probably is evidence for a considerable amount of outgassing of the interior and high internal temperature.

This research was initiated and concluded during summer meetings of the "TYCHO" Study Group under the National Aeronautics and Space Administration Contract No. NSR-24-005-047 with the University of Minnesota. The senior author would like to acknowledge support during the intervening period by the Sloan Foundation. Calculations reported herein were performed at the University of Colorado, Princeton University, and the California Institute of Technology. This study represents contribution 0000, Division of Geological Sciences, California Institute of Technology.

MODEL CONSTRUCTION

We know the diameters and masses and hence the mean densities of most of the planets and satellites in the solar system. We also know the surface temperatures of most of these bodies. For some of these objects we also know the flattening, either from optical or dynamical measurements, and this is related to the internal distribution of mass of the object. Even this limited amount of information is sufficient to make useful comparisons between the internal composition and structure of the various planets. Other information is also available which is not unrelated to the interiors of the planets; for example, the presence or absence of an atmosphere, the presence or absence of a magnetic field, the optical properties of the surface, and in some cases, direct observation of the surface features. This additional information provides clues more than strong constraints. From the study of orbits we can infer the Q, or tidal friction, of some of the planets and satellites.

We know very much more than this about the interior of the Earth. From seismic measurements we have the internal structure. Geology, geophysics and geochemistry combine to give information relevant to the evolution of the Earth. Assuming that the Earth is a representative planet of the terrestrial type it seems reasonable to use the Earth as a starting point in discussions of planetary interiors and possible evolutions and to test the known properties of the other planets against predictions made on this basis for planets of corresponding mass. The first step in such a program is to determine the mean densities of Earth-type bodies of different sizes. This calculation requires an equation of state which we can get from the Earth by knowing its mass, moment of inertia and periods of free oscillation (ref. 2). Kovach and Anderson (ref. 3), using this equation of state, designed planetary models and obtained mean density as a function of total mass. They assumed that the transition region in the upper mantle of the Earth was due to pressure induced phase changes and that the core was an iron-rich alloy, chemically distinct from the material of the mantle. If we knew only the mean densities of the planets, Venus and Mars would survive this first test. In other words both Venus and Mars could be considered smaller versions of the Earth, all having roughly the same composition and structure. The Moon is much lighter and Mercury is much heavier than the corresponding Earth-like bodies and this is most easily interpreted as variations in the iron content. The Moon has a density almost identical to that of the Earth's mantle, decompressed to the appropriate pressure. Although the uncertainties in the radius of Mercury are large it is apparently an iron rich planet, roughly the composition of the Earth's core.

The moment of inertia of Mars indicates that it is a much more homogeneous body than the Earth. Although its mean density implies an iron content similar to that of the Earth, it must be more evenly distributed throughout the planet rather than being concentrated at the center in the form of a core. Kovach and Anderson showed that the mass and moment of inertia of Mars could be satisfied by making it a homogeneous self-compressed body having an overall composition similar to the Earth. More recent determinations of the mean density of Mars suggest that it has a slightly smaller density than given by the above procedure. This can be interpreted in three ways:

1) Mars has a slightly lower iron/silicon ratio than the Earth,
2) More of the iron in Mars is oxidized than in the Earth, i.e. more of the iron in the Earth has been reduced, either in the process of core formation or as an initial condition of the preplanetary material.
3) Mars has retained more light volatiles than the Earth.

Since we do not know the moment of inertia of Venus, a similar test cannot be applied. On the basis of observational data, Venus may or may not have a core. However, the presence of a thick atmosphere on Venus is suggestive of outgassing and, therefore, differentiation and a metallic core. This is consistent with thermal history calculations to be presented later. Neither Mars no Venus seems to have an appreciable magnetic field; this is consistent with the absence of a core on the one hand and a slow rotation rate on the other.

Anderson and Kovach (ref. 4) extended their earlier study by calculating the mean densities of different sized homogeneous planets with different compositions, (See Figure 1). Their index of composition was the mean atomic weight which for the Earth is about 30. The resulting calculation gives the density versus depth for various size bodies of a given composition. Homogeneous bodies constructed in this way will be the starting point of the present computations. These "protoplanets" can be viewed as homogeneous

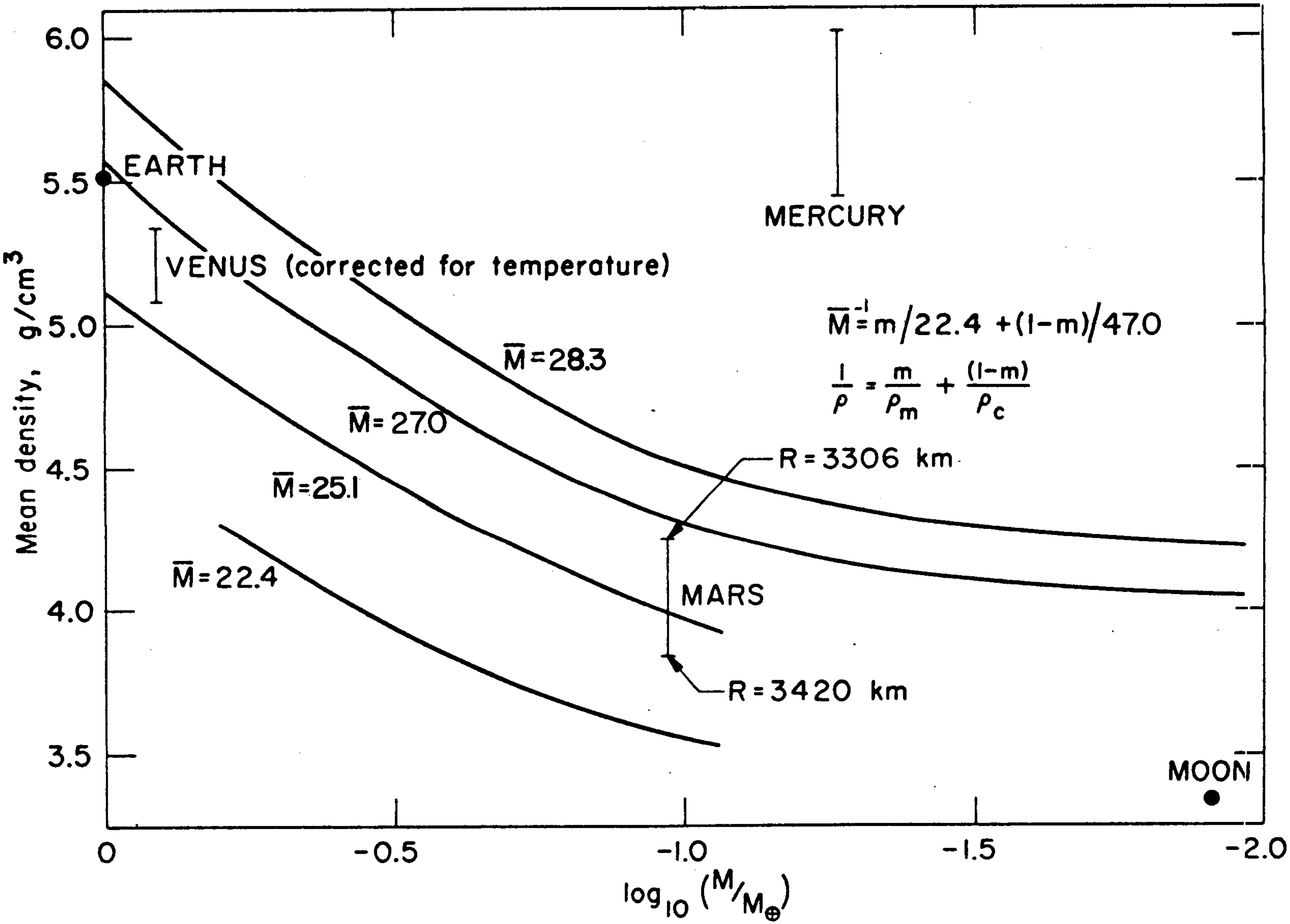

Figure 1 Mean density versus mass, relative to Earth, for homogeneous self-compressed planets of different mean atomic weight, $\overline{M}$. The mean atomic weight of the Earth is about 27.0. These curves represent the density-mass relationship for undifferentiated planets of different composition.

mixtures of heterogeneous particles, in particular, an intimate mixture of iron and silicate or oxide particles. The present Earth has very rapid increases in physical properties in the depth interval between about 300 and 800 km. These are attributed to phase changes from loose packed silicates to close packed silicates or oxides. These pressure induced phase changes are included in the density-pressure relationship used to construct the protoplanets. The density increases with depth due to self-compression and at the appropriate pressure by solid-solid phase changes.

We now assume that the heat producing radioactive elements are also homogeneously distributed throughout the initial body. We take the Wasserburg et al (ref. 5) ratios of $K/U = 10^4$ and $Th/U = 3.7$ determined from terrestrial rocks. Arguments have been advanced that these ratios are unaffected by magmatic differentiation. These ratios can be compared with $K/U = 8.15 \times 10^{+4}$ and $Th/U = 4.2$ which are appropriate for chondrites and which have been used in most previous calculations of thermal history. MacDonald (ref. 6) has shown that an average U concentration in the mantle between 4 and 5×10^{-8} will give the observed heat flow for an Earth of initial temperature near $1000^{\circ}C$. We will assume that the present U content of the crust and mantle is 4.5×10^{-8} and that the present core is free of radioactivity. In the homogeneous protoplanet the average radioactive abundances throughout the planet are reduced from the values given because of dilution by the metallic core which is assumed to be free of radioactive elements. The age of the larger solid objects in the solar system is taken as 4.5×10^9 years. The known half-lives of the long-lived radio-isotopes are then used to compute the abundances of the starting protoplanet.

We now have established the rate at which radiogenic energy is released in nominally protoearth type planets. The rate at which the body will heat up depends also on the specific heat and the mechanism of heat escape to the surface. We follow Lubimova in assuming a lattice conductivity that varies as $T^{-5/4}$ and a radiative conductivity that varies as $\frac{\sigma T^3}{\epsilon}$ where T is the absolute temperature, σ is Stefan-Boltzmann constant and ϵ is the opacity. The specific heat is taken as 1.2×10^7 ergs/gm and constant.

We take the initial temperature gradient to be adiabatic, i.e. that due to self-compression alone. We than investigated the subsequent thermal history of various sized bodies with several assumptions regarding the surface temperature, and lattice conductivity. MacDonald (ref. 6) presents calculations which show the effect of changing the opacity.

INITIAL TEMPERATURES

It is necessary to estimate the initial temperatures in the protoplanets before proceeding with the thermal evolution problem. Recent reviews of this problem are given by Lubimova (ref. 7) and MacDonald (ref. 8). According to current ideas, the Earth and terrestrial planets were formed by accretion of particles from a gas-dust protoplanetary cloud. Although the mechanism of accretion remains poorly understood, it is still possible to place crude constraints on the end product. Most discussions, including the present one, assume that accretion required from 10^6 to about 3×10^8 years. Five possible contributions to the initial temperature have been recognized, and are summarized below:

1) Short-lived radioactivities, with half-lives less than 10^8 years could have produced substantial heating if the planets formed shortly after nucleosynthesis. The lack of radiogenic Xe^{129}, due to decay of I^{129} ($T_{1/2}=1.6 \times 10^7$ years) on the Earth seems to require an interval greater than about 5×10^7 years between nucleosythesis and the start of accretion. We can, therefore, eliminate from consideration initial heat from this source.

2) Long-lived radioactivities, such as K^{40}, U^{235}, U^{238}, and Th^{232}, can play a minor part, if the accretion time is sufficiently long.

3) Solar radiation leads to equilibrium black body temperatures in the range $200^{\circ}K$ - $600^{\circ}K$, depending on the planet under discussion and any postulated variations of solar brightness in the past. This gives a probable minimum temperature, unless special assumptions are made about the primitive solar nebula.

4) Adiabatic compression of the interior produces a temperature increment which can be reasonably estimated, and which, added to 3) supplies a minimum starting temperature.

5) Gravitational energy in the protoplanetary dust cloud must be dissipated or stored in some other form. Impact of accreting matter with the growing planet turns all the energy into heat, most of which is reradiated into space as black-body radiation. The T^4 dependence of black-body radiation insures a stable situation in which the rate of input of gravitational energy is balanced by the radiation rate:

$$\frac{g(r)\,M(r)}{r}\,\frac{dr}{dt} = \sigma T^4 \qquad (1)$$

The gravitational term on the left should be reduced by the amount of energy required to compress the planet and the amount required to heat the incoming material to the equilibrium temperature T. Energy may be stored or released chemically due to any postulated mineralogical changes occurring on impact. One can easily show that all these effects involve energies small compared with the initial gravitational energy. The above equation is then appropriate to determine the surface temperature at any stage of accretion, and one can then show that the total amount of energy which goes into heating the planet is less than 10% of the gravitational energy of the dust cloud, the remainder being lost by radiation.

If the Earth and Moon accreted at a constant rate in 10^8 years the equilibrium surface temperatures would be of the order of $250^{\circ}K$ and $73^{\circ}K$

respectively. An Earth which accreted uniformly in 10^9 years would have a temperature of 145^oK. Using Ter Haar's (ref. 9) estimates of an accretion rate which accelerates with time, the equilibrium surface temperatures for the Earth, Mars, and the Moon will be, respectively, 1100^oK, 330^oK, and 110^oK.

One can assume some accretion history and estimate the temperature profile due to the input of gravitational energy. Most authors, using similar assumptions, deduce a temperature maximum near the surface. Combining this with the adiabatic gradient, one can produce an initial temperature which increases in the outer part of the planet and remains roughly isothermal elsewhere (ref. 7, 8). This average temperature is of the order of 1000^o, 300^o, and 100^o for the Earth, Mars, and the Moon, respectively, In this paper, we study models in which the role of the gravitational energy is minimized, and the more easily estimated adiabatic and solar temperature contributions are taken. An isothermal interior can only be formed by putting in the near-surface maximum due to gravitational energy. We feel that there is some merit in studying the consequences of a distinctly different model without this poorly known contribution. Our starting temperatures are therefore taken as adiabatically increasing with depth. The surface temperature is a parameter of interest, whose value is given by the solar equilibrium temperature plus a contribution.

RESULTS OF CALCULATIONS

Figure 2 gives the development of the internal temperatures for an undifferentiated planet having the mass of the Earth and an initial surface temperature of 330^oK. The calculations were performed for an opacity of $100\ cm^{-1}$ and for two values of the lattice conductivity. The lower value is typical of crustal rocks and the higher value is typical of an iron rich rock which presumably is more representative of the implied early conditions of

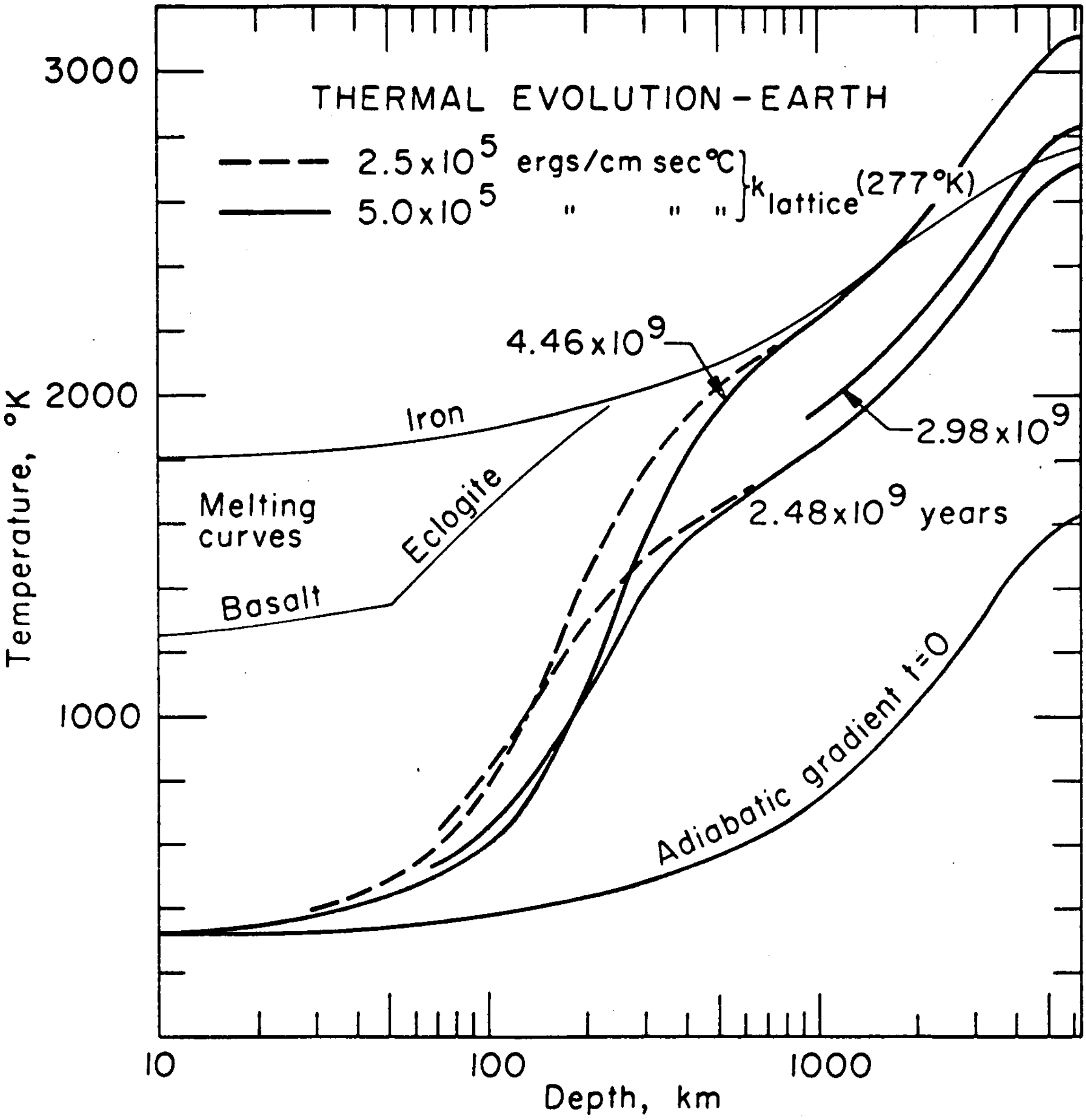

Figure 2 Initial temperature and temperatures at three later times for an undifferentiated Earth with a relatively cold origin.

an undifferentiated planet. The melting curves are from Strong (ref. 10), and Yoder and Tilley (ref. 11). The calculations ignore latent heats of fusion and the possibility of heat transport by convection and therefore represent maximum temperatures obtainable under the assumptions stated. The melting temperature of iron is reached at the center of the Earth at about 2.7×10^9 years after formation and is exceeded at the present time, taken as 4.46×10^9 years, below some 1500 - 1800 kilometers depth. These conclusions are not altered by reasonable variations in the lattice thermal conductivity. This is partially due to the dominance of the radiative term in the deep interior and at high temperatures.

The present depth to the core is 2898 km and it presumably contains most of the free iron in the Earth. The core of the Earth is thought to contain iron, silicon, and nickel.

Any free iron present below about 1500 km will melt and drain downward enriching the central region in iron and depleting the region above.

At the onset of melting, however, the latent heats, heat transfer by convection, rearrangement of radioactivities and the increase in gravitational energy must all be taken into account in the further thermal evolution. These calculations are only for the purpose of predicting whether or not melting will commence. The conclusion at this point is simply that an initially adiabatic, homogeneous Earth-sized planet, even with a starting surface temperature as low as $330^{\circ}K$ will exceed the melting point of iron and the melting point is first exceeded near the center of the planet. Previous thermal history calculations, using an initially isothermal planet, have indicated that the melting point of iron is first exceeded high in the mantle leading to a molten iron layer which becomes unstable and drops catastropically toward the center, releasing enough gravitational energy to melt and separate the remaining iron. Our calculations would have the core growing more slowly and the gravitational energy added gradually, at least in the earliest stages of core formation. The end result, however, for an initially homogeneous Earth-sized body containing free iron is the same; namely, the eventual formation of a central iron-rich molten core. The lighter, lower melting point silicates, and volatiles would escape toward the surface in the process of core formation. The amount of gravitational energy released by core formation is, of course, independent of the mechanism and the discussion of Birch (ref. 12) is relevant in any case.

We consider the existence of a crust, a core, a hydrosphere, and an atmosphere all to be evidence of a differentiated, outgassed planet. Volcanism and tectonism are current manifestations of present differentiation and readjustments to past differentiation.

Note that the melting point of silicates is not reached in the present calculation. In the process of core formation approximately 2.5×10^{10} erg/g of gravitational energy is transformed to thermal energy and this is equivalent to an average rise of temperature of about 2000° (ref. 12). The temperature profile in the interior of a planet in the process of differentiation will be approximately that of the lowest melting component, or the minimum eutectic point of the multicomponent system. This temperature gradient will presumably be controlled by silicates in the upper mantle and iron in the lower mantle and core. If the starting surface temperature of the Earth is closer to $1000^{\circ}K$ than $300^{\circ}K$, the process of melting and differentiation will already be under way while the planet is still accumulating material.

We previously estimated that $330^{\circ}K$ is an approximate initial surface temperature for Mars. Figure 3 shows the temperature distribution after 4.46×10^{9} years for the Earth and for Mars for this initial temperature.

Although the temperature near the center of the Earth exceeds the melting point of iron, identical assumptions for a Mars-sized body give present day temperatures that are everywhere below the melting point of iron. Free

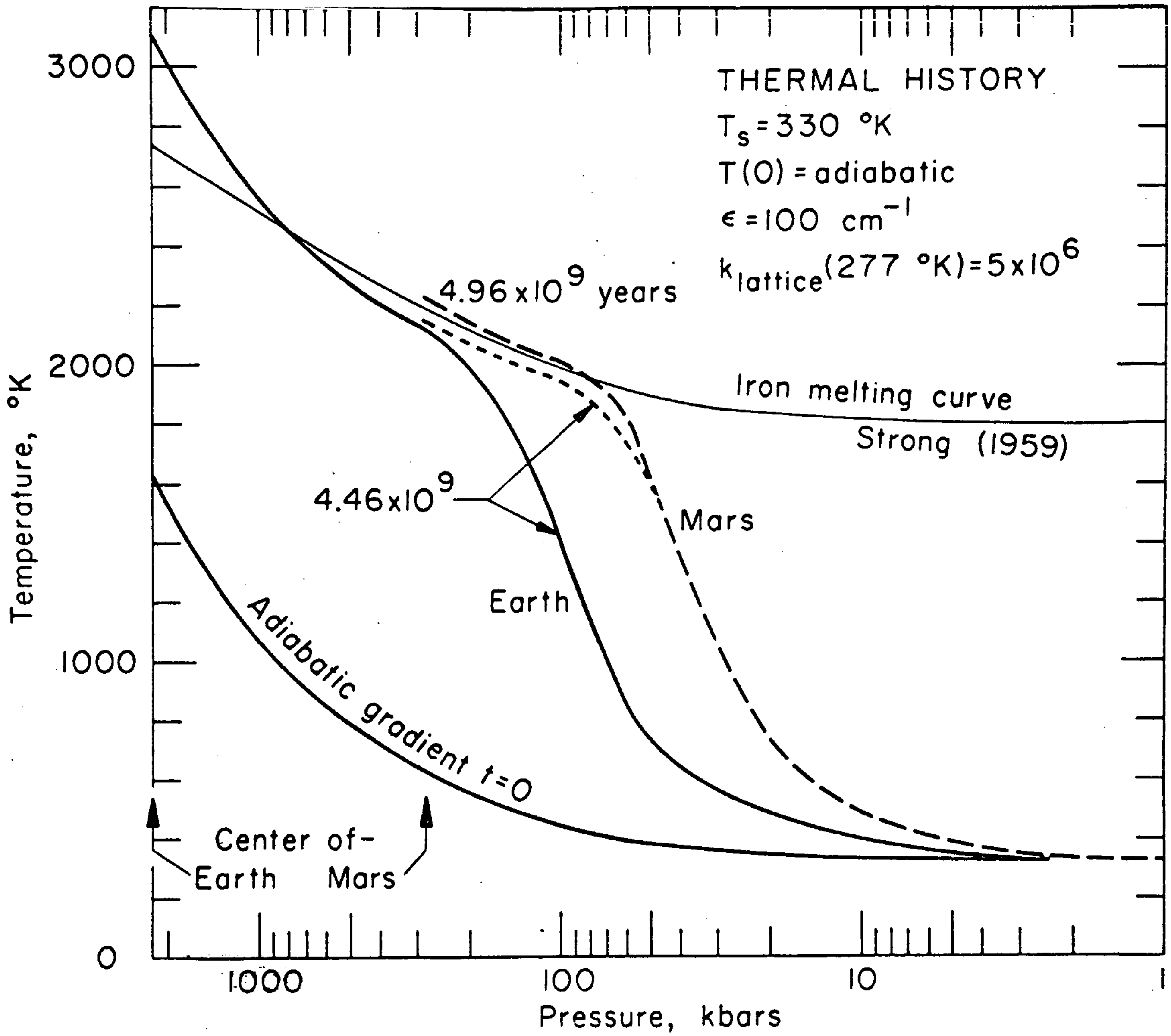

Figure 3 Initial and present temperatures for Mars and the Earth having an initial surface temperature of 330° K and an initially adiabatic thermal gradient. Also shown are temperatures for Mars at 4.96 x 10^9 years. Note the presence of a liquid core in the Earth and the absence of melting in Mars.

iron in the interior of Mars will be solid and the process of core formation presumably has not started. However, the temperatures below about 500 km are very close to the melting point; and if the viscosity of the silicate phase is low enough, a certain amount of separation in the solid phase may be possible. However, this is a much slower process than the separation in the liquid phase. The thermal history calculations are, therefore, consistent with previous calculations regarding the homogeneity of Mars--present-day temperatures are too low to cause core formation which presumably is the trigger for the later events associated with differentiation. Temperatures in the interior are, however, high enough to be consistent with a certain amount of outgassing. Temperatures are also high enough to suggest that the process of core formation will begin within 0.5×10^9 years.

Figure 4 gives the sequence of events for Mars and Earth for a starting temperature of $400^{\circ}K$. In this case the melting point of iron is reached at the center of the Earth about 2×10^9 years after planetary formation and the present ($t = 4.46 \times 10^9$ years) temperatures at the center of Mars (not shown) are just about at the melting point of iron. This figure also shows another extrapolation of the melting point of iron which probably represents an upper bound but does indicate another source of uncertainty in the conclusions based on thermal history calculations. The curve of Strong (ref. 10) is closer to recent estimates of the melting point of iron and will be the curve shown in subsequent figures.

Figure 5 gives the thermal history calculations for a very low starting surface temperature, $100^{\circ}K$. In this case, neither Mars nor the Earth will have reached the melting point of iron and both will be undifferentiated planets if they started as homogeneous bodies.

The evolution of Mars with a starting surface temperature of $330^{\circ}K$ and the two different lattice conductivities is given in Figure 6. No melting of iron occurs and this conclusion does not depend on the choice of lattice

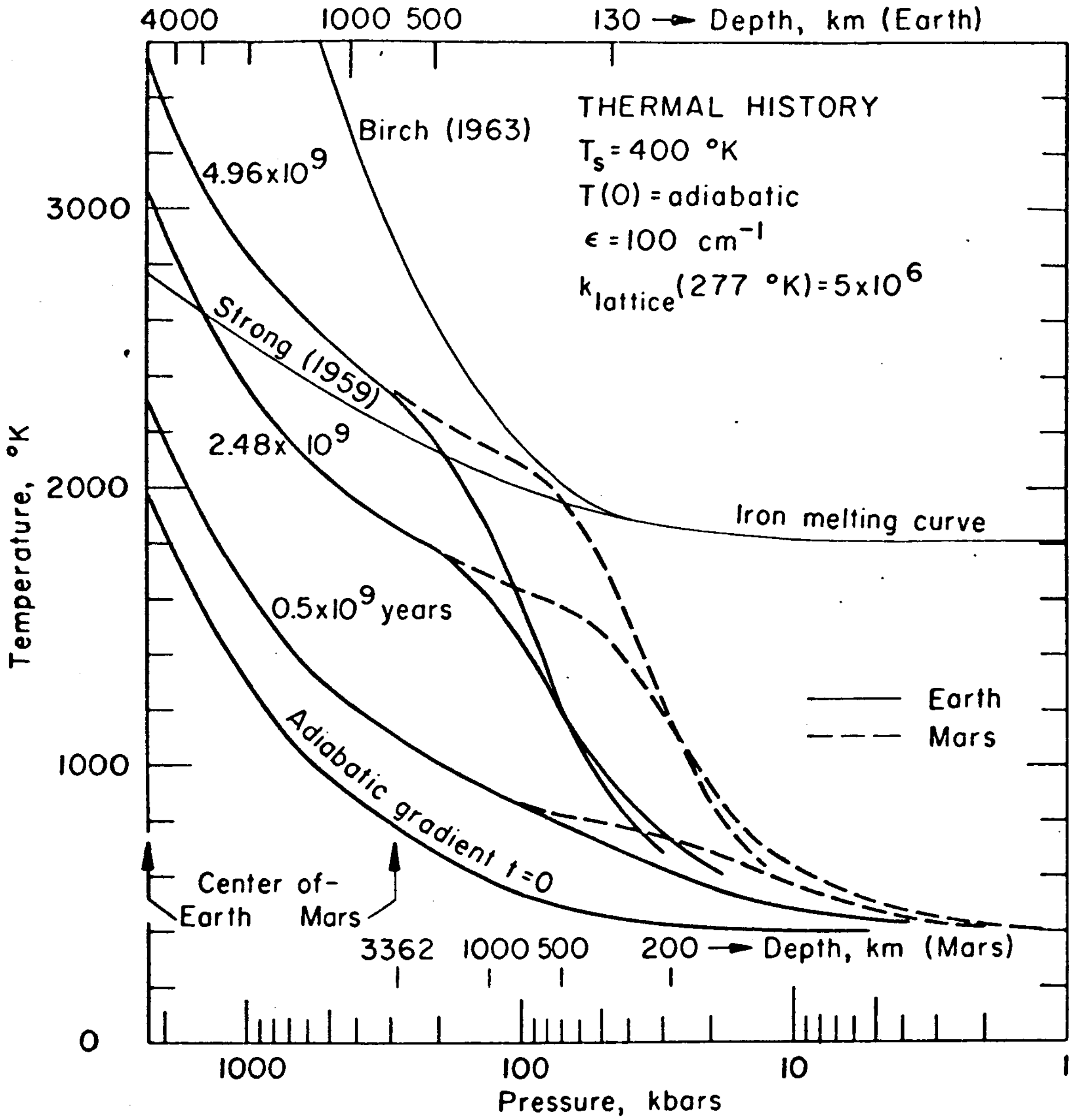

Figure 4 Thermal evolution for Mars and Earth for a starting surface temperature of 400°K.

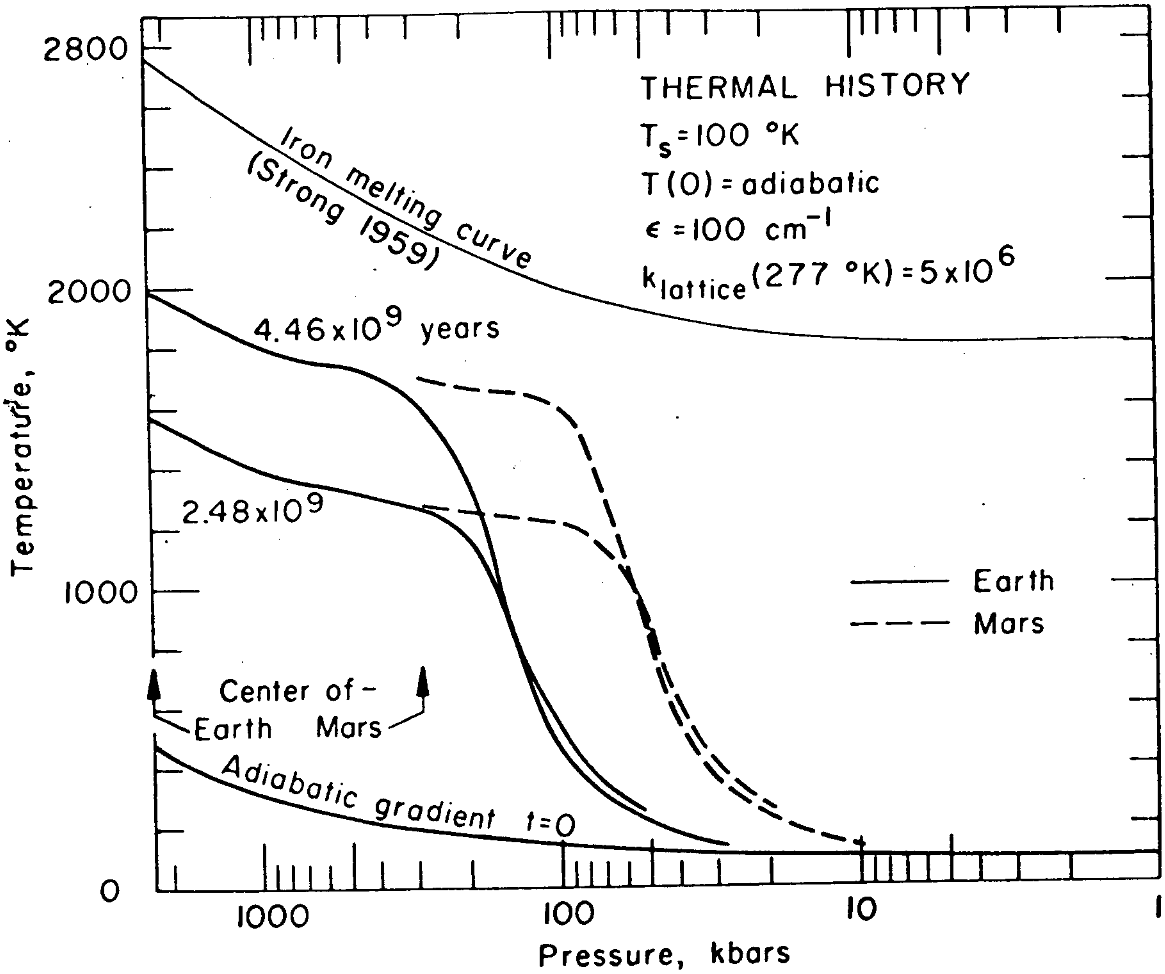

Figure 5 Thermal evolution for Mars and Earth for a starting surface temperature of 100° K.

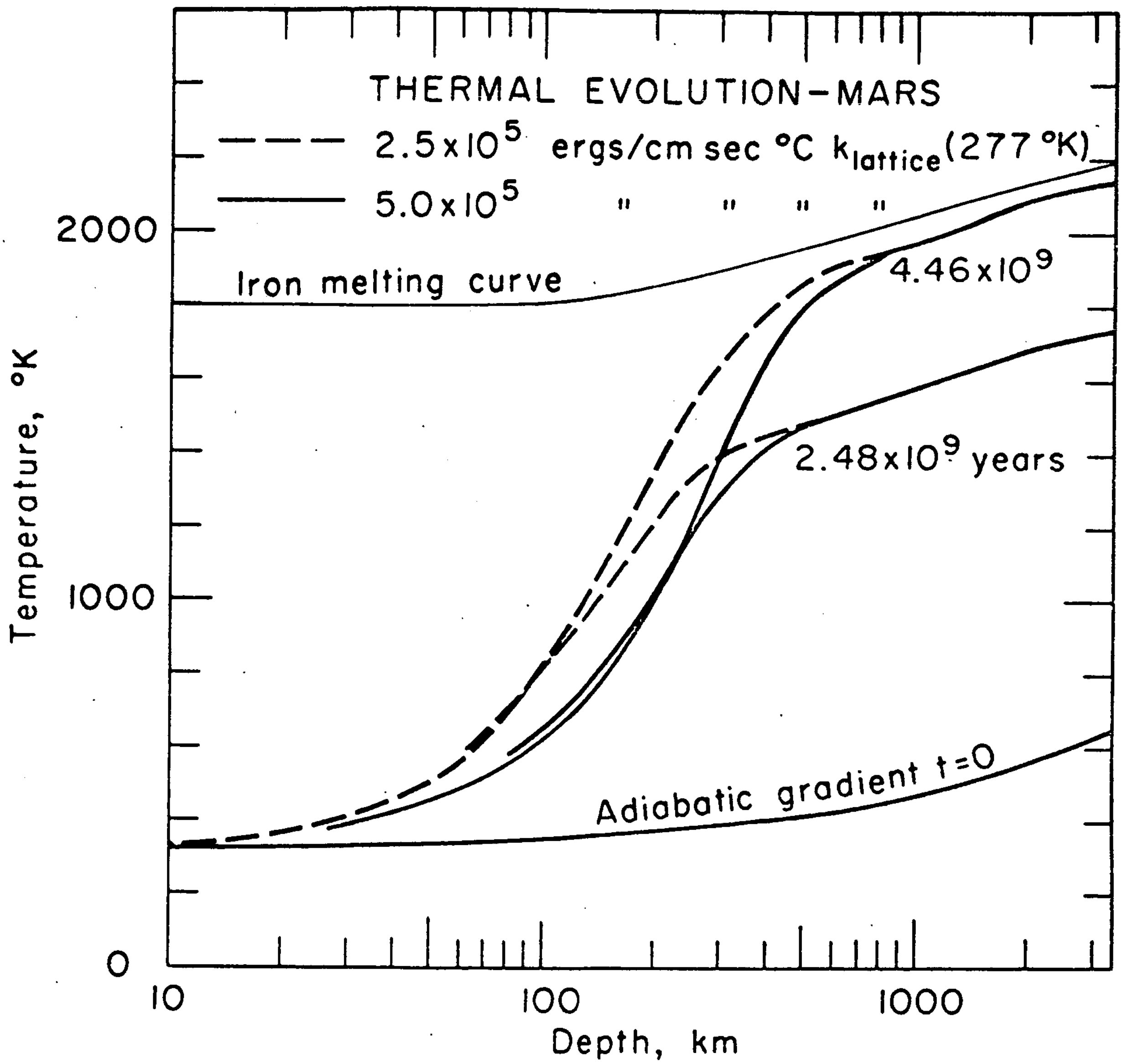

Figure 6 Thermal evolution for Mars having a starting surface temperature of 330°K illustrating the effect of lattice conductivity.

conductivity unless it is much lower than 2.5×10^5 ergs/cm sec^{o}C.

So far we have only been comparing Earth and Mars. The conclusion is that it is fairly easy to differentiate an Earth but difficult to differentiate a Mars if our estimates of starting temperatures and radioactive abundances are close to the true conditions.

For any given size planet there is a critical starting temperature which will give a present central temperature high enough to melt iron, Figure 7. Or, put another way, for a given starting surface temperature there is a minimum sized planet which will be able to begin melting its iron in the lifetime of the solar system. For an initial temperature of 330^{o}K this critical size planet falls between the masses of Mars and the Earth, roughly about 2×10^{27} grams. If the initial temperature is related to mass approximately as discussed in the preceding section, then the Earth is represented by a point at T_s = 1000^{o}K, and Mars by a point near T_s = 330^{o}K. Hypothetical intermediate planets would be represented by a continuous curve between these points. Mars then appears to have such a mass that its center is now about at the melting point of iron. Venus, with a mass of $.5 \times 10^{28}$ g, must have melted its iron, given the postulates of our model. This is, of course, compatible with the inference that Venus is differentiated, based on the existence and composition of its thick atmosphere.

A Moon-sized body with the same composition as the Earth is below critical mass and will not differentiate. However, the density of the Moon alone tells us that it is not the same composition as the Earth and is probably deficient in iron relative to the Earth as a whole. The average concentration of radioactive elements in the Moon is probably like that in the Earth's mantle. The uncompressed density of the Moon is remarkably close to the uncompressed density of the Earth's mantle. The simplest explanation is that the Fe/Si ratios are the same for the Moon and the mantle and quite different from the Fe/Si ratio of Mars, Venus, and the Earth taken as a whole. Urey has

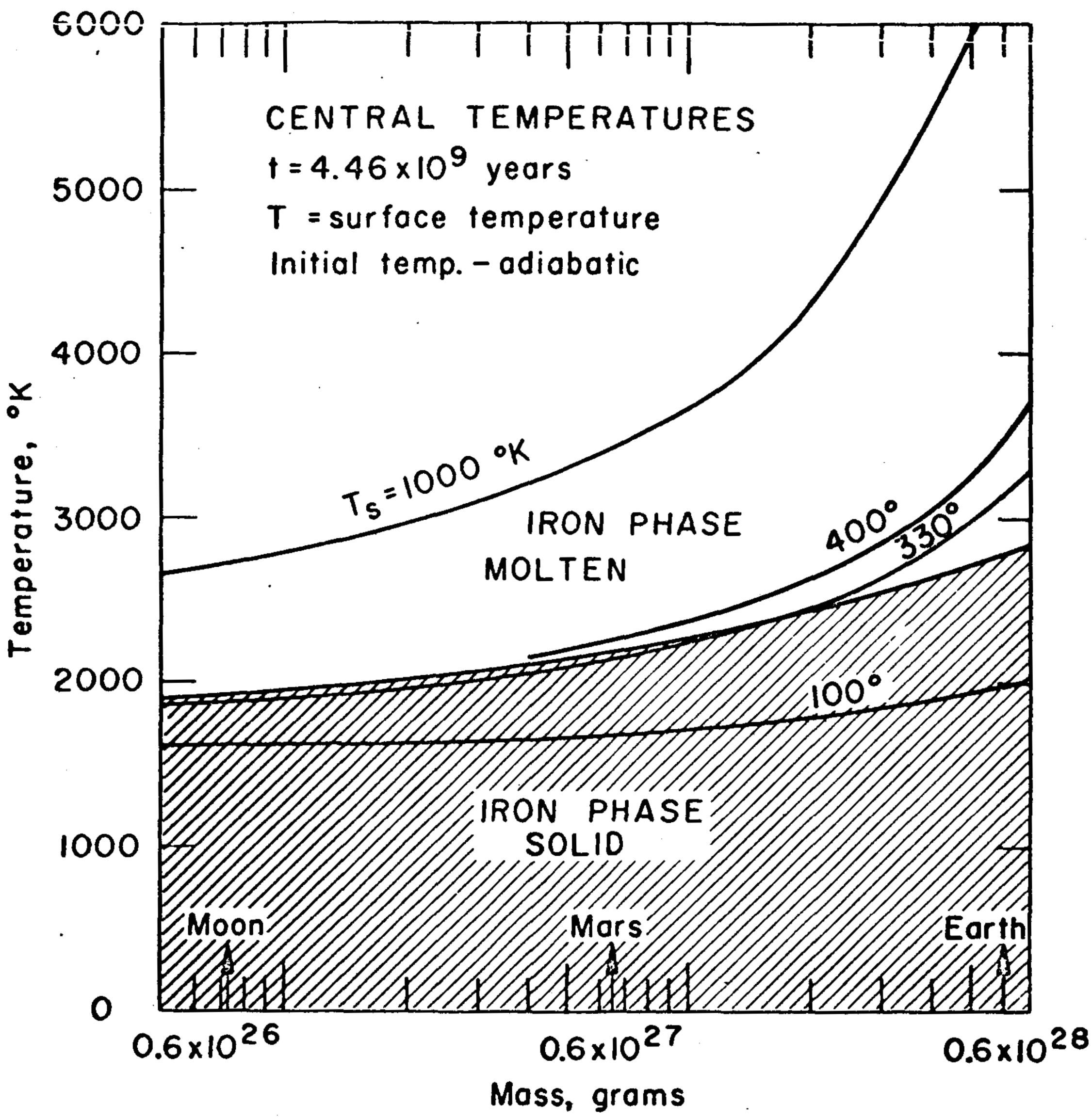

Figure 7 Central temperature (also maximum temperature) at $t = 4.46 \times 10^9$ years as a function of planetary mass for various initial surface temperatures. The initial temperature gradient is adiabatic and the planets all have the same composition as the present Earth. The hatched region represents planets that are completely solid, assuming that iron is the minimum melting phase. The boundary between the hatched and unhatched regions represents the melting point of iron at the pressure at the center of the planet. The label "iron phase molten" refers to conditions near the center of the planet. In general, the isothermals are not loci of physically realizable planets.

argued that the Fe/Si ratio of the Sun is similar to that of the Moon and, therefore, the Moon may represent primordial matter.

If one starts with an isothermal interior, then our principal conclusions are uneffected. The iron will begin melting in a shell instead of at the center. The conclusions follow basically from our relation between mass and average initial temperatures and from the radioactive concentrations obtained by mixing iron into mantle material to form a protoplanet.

TEMPERATURES IN THE MOON

It is obvious from Figure 1 that the mean density of the Moon is much lower than the other terrestrial planets. It is, in fact, very close to the density of a material having a mean atomic weight of about 22.4 which is the mean atomic weight of the mantle (ref. 13, 14). The most reasonable inference is that the composition of the Moon and the Earth's mantle are very similar, at least in the major elements. In the absence of information to the contrary, it is reasonable to assume that the concentration of radioactive elements in the Moon is also similar to that in the mantle rather than that in the Earth as a whole. An alternate assumption is that the radioactivity of the Moon is similar to chondritic meteorites. Thermal history calculations for both assumptions are shown in Figure 8. In both cases the melting point of silicates is exceeded early in the history of the Moon. In these calculations the initial temperature is taken as $273^{o}K$ and constant and the opacity is 100 cm^{-1}.

In the chondritic model melting commences at about 2.1×10^9 years after formation at a depth of 300 km and proceeds rapidly inward. Latent heat of melting and heat transfer by differentiation, ignored in these calculations, will keep the interior from being totally fluid. If heat removal by fluid convection is not allowed a temperature in excess of about 300^{o} of the

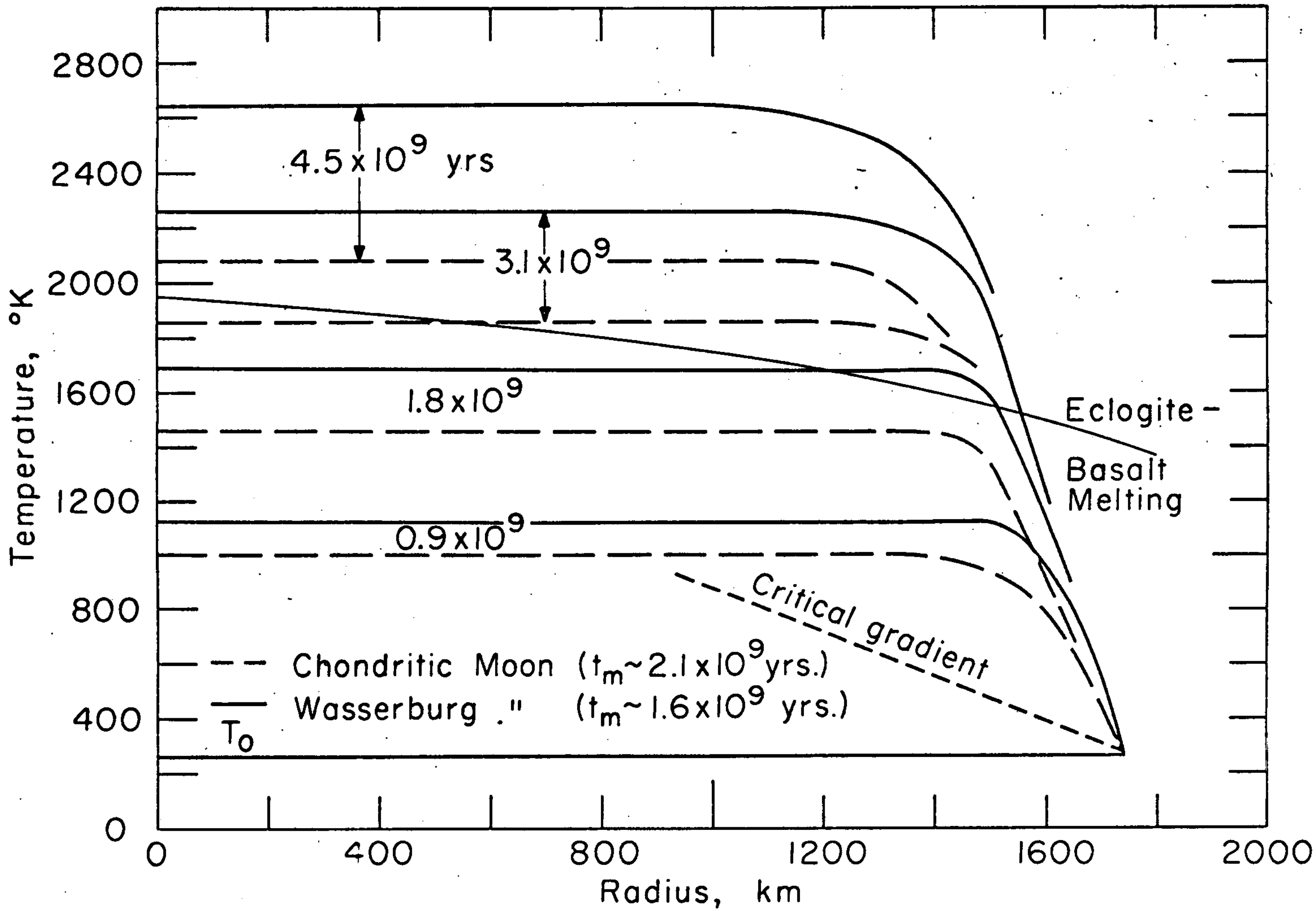

Figure 8 Variation of temperature with radius and time for moons with chondritic and terrestrial (Wasserburg et al - 5) radioactive abundances. t_m is estimated time of initial melting of the silicate phase. Thermal gradients greater than critical gradient imply density decreasing with depth.

melting curve implies total melting and this is achieved throughout most of the interior of the Moon for both assumptions regarding the radioactive content. A 200 km thick solid crust is also implied. If the interior of the Moon is to stabilize at or below the melting point of silicates, extensive volcanism is required to remove the excess heat and this will commence after about 2.1×10^9 years for a chondritic Moon and 1.6×10^9 years for a Moon having terrestrial abundances of radioactivities.

The dashed line labelled "critical gradient" represents the thermal gradient in the Moon that will lead to constant density with depth. The temperature rise in the interior of the Moon which is implied by the thermal history calculations gives supercritical gradients near the surface and this in turn implies a density that decreases with depth.

Figure 9 shows the "present" temperatures for various Moon models ignoring latent heats and convective transfer. The radioactive abundances must be reduced below 2/3 of the terrestrial mantle abundances if melting is to be avoided. It is difficult to avoid having a "hot" Moon without having extensive volcanism in the past. An alternate way to achieve a "cold" Moon is to have its formation antedate that of the Earth by several billion years.

Above each curve is the value for H, the present heat flow, ignoring convection. This is obviously an important quantity to measure on the Moon and, along with measured radioactivities, will place useful constraints on further thermal history calculations and discussions of the composition and evolution of the Moon.

The present calculations suggest that the Moon is a differentiated body, presumably basalt over something like pyrolite. We would expect surface concentrations of uranium more like that of basalt, $\sim 6 \times 10^{-7}$ gg^{-1} than that of ultrabasic rocks, $\sim .01$ to 2.5×10^{-7} gg^{-1}.

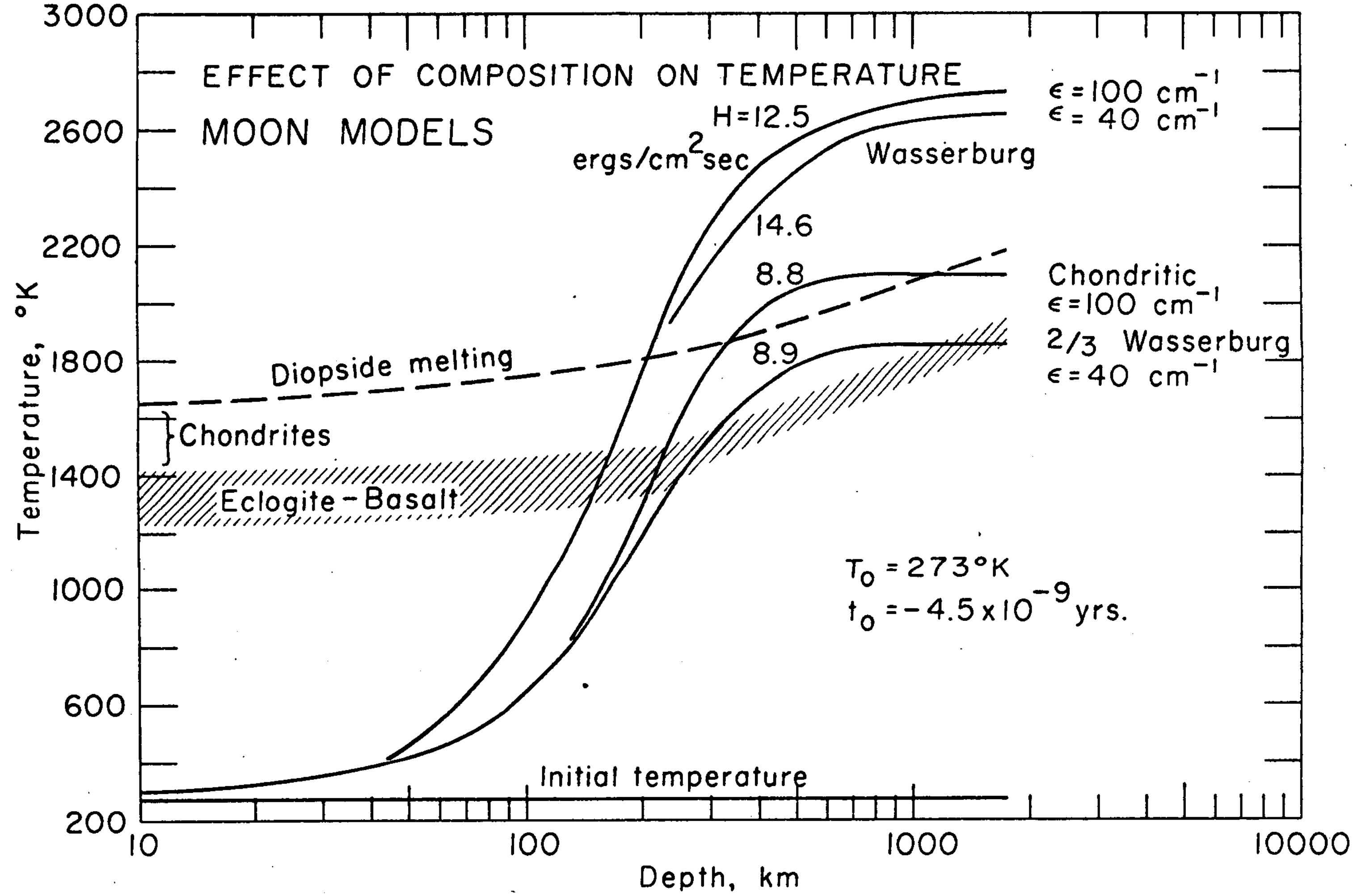

Figure 9 Present temperatures for various Moon models. The parameter H is the present heat flow predicted for these models.

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